BOTTOM WATER FORMATIONS IN THE NORWEGIAN SEA

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by

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ABSTRACT

The deep basins of the world oceans are filled with waters of very low temperature. A fundamental problem of oceanography concerns the nature of the sources of that water: their locations and temporal variations of strength. A stability analysis of temperature and salinity data recently taken in the Norwegian sea shows that bottom water probably does not form in that area.

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TABLE OF SYMBOLS

c	speed of sound in sea water (m/sec)
E, E_h, E^*, E_r	stability parameters
g	gravitational acceleration
h	depth beneath ocean surface (m)
p	pressure
s	salinity (parts per thousand)
T	temperature (degrees centigrade)
v	downward velocity of water mass
ρ	density of sea water (gm/cm^3)
E_r	error in stability parameter
$(\ )_A$	adiabatic process

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1. Introduction

The deep basins of most of the world oceans are filled with waters of very low temperature. The source of these cold bottom waters has been the subject of considerable research in oceanography. Nansen, in 1912, formulated what has become the accepted theory for the formation of these waters. His theory, stated briefly, is as follows. During the cold season, the salinity of the surface layer in the high latitudes is increased either through advection of more saline water from lower latitudes or through freezing of the surface water *in situ*. Concurrent with or subsequent to this salinity increase the temperature of the surface water is lowered by heat loss to the atmosphere, through the processes of radiation, convection and conduction. The combination of increased salinity and decreased temperature produces surface water which is more dense than the stratum of water immediately below the surface layer. This more dense water will sink, mixing with the underlying waters and producing a thin layer of homogeneous water. As the process of formation proceeds the thickness of this homogeneous layer will increase, and if enough dense surface water forms the layer will ultimately extend from the surface to the bottom. If the process of dense water formation is not sufficiently intense or if it does not continue for a sufficient length of time the homogeneous layer will not extend to the bottom but will terminate at an intermediate depth and be destroyed by lateral mixing at the conclusion of the surface formation period for the dense water.

Since the formation process for bottom water depends upon interaction between the sea and the atmosphere it is not likely that the

process proceeds at a uniform rate, but rather that it occurs at greater or lesser rates depending on variations in the upward transfer of heat and in the surface salinity. Variations in wind, precipitation, air temperature, etc., should vary the rate of bottom water production over short intervals of time, perhaps of the order of days or even hours. The result would not be a column of water having a uniform density from surface to bottom, but rather a column having vertical density variations corresponding to periods of formation of waters of different density at the surface. During a relatively long period of formation sufficient water of high density might be produced so that a stratum could retain its characteristics long enough to enable it to reach great depths without complete loss of identity through mixing, and thus be detectable as inhomogeneity in the density distribution.

Static vertical stability in a fluid is a measure of the restoring force on a fluid element when it is given a small adiabatic vertical displacement from its undisturbed position. This force is proportional to the difference between the density of the displaced element and the density of its surroundings. In a stably stratified fluid an element displaced downward will be less dense than its surroundings and the resulting buoyant force will tend to return the element to its former location. In a neutrally stratified fluid the displaced element will have the same density as its surroundings after displacement and will experience no buoyant force. In an unstably stratified fluid the element, on being displaced downward, will have a density greater than that of its surroundings and will experience a negative buoyant force tending to displace the element farther from its former location.

In an area of bottom water formation where atmospheric conditions vary with time, the stability of the surface layer should vary from stable through neutral to unstable, depending upon the heat and salt content. After the beginning of formation of actual bottom water (after the initial formation of the surface-to-bottom homogeneous column) the stability of the deeper columns should be essentially neutral. Finite strata of unstable water might be interspersed from time to time, if there have been periods of extensive formation of very dense water.

According to Sverdrup et al. (1942), the favored areas for formation of bottom water in the North Atlantic Ocean are the Labrador, Irminger, and Norwegian Seas. An extensive time series of hydrographic station data is available for Weather Station Ship Mike (Mosby undated). Ship Mike is located in the southeastern Norwegian Sea at latitude 66° N. longitude 02° E. Figure (1) shows the bathymetry of the southern Norwegian Sea and the location of Ship Mike. These data were analyzed, through stability calculations, to determine whether or not bottom water formed at that location and, if so, the temporal variation of the process.

2. The Data

The data used for this investigation consist of 1253 hydrographic stations taken at Ship Mike during the period January 1949 through December 1953 (Mosby undated). It was obtained in machine form (punched cards) directly from the Geophysical Institute, University of Bergen, through the cooperation of Professor Håkon Mosby. Observations were made using four different combinations of depths for the Nansen bottles and reversing thermometers. The four combinations and the number of observations made using each are shown in Table 1.

TABLE 1

SAMPLE DEPTHS FOR THE FOUR TYPES OF OBSERVATIONS TAKEN AT SHIP MIKE AND THE NUMBER OF EACH TYPE OF OBSERVATION TAKEN 1949-1953

Type	a	b	c	d	
Depths (meters)	0	0	0	0	
	10	10	10	50	
	25	25	25	150	
	50	50	50	300	
	75	75	75	600	
	100	100	100	1000	
	150	150	150		
	200	200			
	300	400			
	400	600			
	500	1000			
	600				
	800				
	1000				
	1200				
	1500				
	2000				
	235	391	231	396	Total number of each type

The four combinations are fairly evenly distributed among the months. Occasionally a month passed with no type (a) observation but

for most months there is at least one and are as many as five observations at depths greater than 1000 meters.

Mosby (1959) has carefully evaluated the accuracies of the observations. In general, standard oceanographic accuracies in temperature, salinity and depth were achieved. However, four to five percent of the salinity determinations contained errors greater than 0.01 parts per thousand. Observations having this reduced accuracy are not identified in the data. There are position errors in the location of the observations, due presumably to the inability of the observing vessel to maintain station in heavy weather. Most of the observations were taken within 15 to 20 nautical miles of the assigned station. With very few exceptions the observations were made in the morning hours, at about 1000 local time.

3. The Stability Calculations

Nansen's theory for the formation of bottom water has been previously discussed. A primary requirement for such formation is the presence of a column of water, extending from the surface or near surface layer to the ocean floor, which exhibits small stability. That is, the column must have a nearly uniform density from top to bottom, exclusive of density changes due to adiabatic heating. The calculation of vertical stability was thus selected as the means for determining the occurrence of conditions suitable for the formation of bottom water at Station Mike. Vertical stability can be calculated using the formula of Hesselberg and Sverdrup (Sverdrup et al. 1942):

$$E = \frac{\partial \rho}{\partial S}\frac{dS}{dh} + \frac{\partial \rho}{\partial T}\left[\frac{dT}{dh} - \left(\frac{dT}{dh}\right)_A\right] \qquad (1)$$

where ρ is density in situ, S is salinity (parts per thousand), T is temperature, h is depth, and $(dT/dh)_A$ represents the adiabatic temperature change with depth. Here, E is the net restoring force per unit volume and displacement, divided by the gravitational acceleration, g. This equation is inconvenient for digital computer calculations. The terms dS/dh and dT/dh are evaluated directly from the data. The other terms are evaluated using extensive tabulated data. The inclusion of these tables lengthens the calculation time and reduces the amount of the memory core available for output data storage during the calculation process. The stability parameters developed by Pollack (1954) are much more convenient for computer use and were chosen for this investigation.

Pollack's formulas are divided into three families. Calculations

made with the first family give stabilities in terms of force per unit mass per unit displacement, denoted by E_i. Those made with the second family give stabilities in terms of potential energy per unit mass per unit displacement, E_i^*. Results using the third family give stabilities as a dimensionless ratio between the <u>in situ</u> density gradient and the adiabatic density gradient, $E_{r(i)}$. The subscript (i) depends upon the method of measuring depth, which may be in terms of pressure (i = p), geopotential (i = G), or geometrical length (i = h). Equations (2), (3), and (4) are examples of Pollack's formulas.

$$E_h = \frac{g}{\rho}\left[\frac{d\rho}{dh} - \left(\frac{d\rho}{dh}\right)_A\right] \tag{2}$$

$$E_h^* = \frac{g}{2\rho}\left[\frac{d\rho}{dh} - \left(\frac{d\rho}{dh}\right)_A\right] \tag{3}$$

$$E_{r(h)} = \frac{d\rho}{dh} \Big/ \left(\frac{d\rho}{dh}\right)_A \tag{4}$$

Here, $(d\rho/dh)_A$ is the adiabatic density gradient. Since the speed of sound is given by $c = \sqrt{(dp/d\rho)_A}$, the above equations can be written in the more convenient forms:

$$E_h = \frac{g}{\rho}\left[\frac{d\rho}{dh} - \frac{\rho g}{c^2}\right] \tag{5}$$

$$E_h^* = \frac{g}{2\rho}\left[\frac{d\rho}{dh} - \frac{\rho g}{c^2}\right] \tag{6}$$

$$E_{r(h)} = \frac{c^2}{\rho g}\frac{d\rho}{dh}. \tag{7}$$

The dimensionless stability parameter, $E_{r(h)}$, was selected for the stability calculations, and will hereafter be written as E_r. Values of E_r greater than unity indicate stability, values equal to unity indicate neutral stability, and values less than unity indicate instability. Negative values of E_r indicate a decrease in density in situ with depth.

Following Pollack, E is related to E_h by

$$E = \frac{\rho}{g} E_h \tag{8}$$

and to E_r by

$$E = \frac{\rho g}{c^2} (E_r - 1). \tag{9}$$

A fairly accurate and simple approximation to equation (9) can be obtained. Assuming that ρ = 1gm/cm^3 , g = 10^3 cm/sec^2 and $c^2 = 2 \times 10^{10}$ cm^2/sec^2, equation (9) can be written as

$$E \doteq \frac{E_r - 1}{2} \times 10^{-7}. \tag{10}$$

Stabilities calculated using Pollack's equation compare closely with those calculated using the Hesselberg-Sverdrup equation. Table 2 illustrates stability calculations made using several of the equations described above. The last two columns have been added to a similar table given by Pollack. The degree of approximation resulting from the use of equation (10) can be determined by comparison of the fourth and seventh columns of the table.

TABLE 2

COMPARATIVE VALUES OF VARIOUS STABILITY PARAMETERS ON A SINGLE SET OF DATA (Note: we here deviate from standard oceanographic practice by using cgs units in E, E_h, and E_r)

Depth m	Temp. °C	Salinity Ppt	$10^{10}E$ Sec^{-2}	$10^{7}E_h$ Sec^{-2}	E_r	$10^{10}E$ from eq. (10) gm/cm^4
0	19.20	36.87				
			-440	-434	.05	-475
10	19.31	36.85				
			-150	-167	.67	-160
25	19.34	36.83				
			- 13	- 15	.96	- 20
50	19.24	36.79				
			610	604	2.33	665
75	18.65	36.79				
			390	364	1.85	425
100	18.24	36.78				
			38*	64	1.04	20
150	17.50	36.56				
			270	242	1.55	280
200	16.45	36.40				
			160	160	1.30	150
300	14.52	36.02				
			120	127	1.25	125
400	13.08	35.77				
			150	155	1.34	170
500	11.85	35.64				
			130	134	1.29	145
600	10.80	35.54				
			100	107	1.23	115
800	9.09	35.39				
			89	96	1.21	105
1000	8.01	35.37				
			84	86	1.20	100
1200	7.27	35.42				
			48	50	1.12	60
1400	6.40	35.35				
			26*	29	1.07	35
2000	4.52	35.5				
			11	14	1.03	15
3000	2.84	34.92				
			8	11	1.01	5
4000	2.43	34.90				
			1	3	.99	-5
5000	2.49	34.90				

*Pollack recomputed these values using the formulas and tables of Hesselberg and Sverdrup

4. <u>The Calculation of Variables</u>

Calculations of E_r using equation (7) require values of both the speed of sound in sea water and the density of sea water *in situ*. These values are normally obtained from tabulated data (e.g. LaFond 1951). Empirical formulas for the speed of sound (Kinsler and Frey 1962) and for the density *in situ* (Eckart 1958) are available. These formulas eliminate the need for tabulated data and are thus more convenient for use in computer calculations.

The Kinsler-Frey formula for the speed of sound is

$$c = 1449 + 4.6T - 0.055T^2 + 0.0003T^3 + (1.39 - 0.012T)(S - 35.0) + 0.017h \tag{11}$$

where c is in m/sec, T in degrees centigrade, S in parts per thousand and h in meters. Results obtained using equation (11) are accurate to within one meter per second.

Eckart's formula for the calculation of density *in situ* is

$$\begin{aligned} \Delta &= 10^4\left(\frac{\rho - 1}{\rho}\right) \\ 10^{-6}K &= 17.795 + 0.1125T - 0.000745T^2 - (0.0380 + 0.0001T)S \\ p_o &= 5890 + 38T - 0.375T^2 + 3S \\ \Delta_o &= 3020 \\ K &= (p + p_o)(\Delta_o - \Delta). \end{aligned} \tag{12}$$

Here, ρ is measured in gms/cm^3, T in degrees centigrade, S in parts per thousand, while p denotes absolute pressure in atmospheres. In the calculations the approximate expression p = 0.101 h was used. This approximation introduces an error of about 5 atmospheres at a depth of 5000 meters

and may account for the slight instability in the 4000-5000 meter layer in Table 2.

Eckart's formula, based upon almost all of the experimental determinations of the density of sea water, is valid for the ranges 0 to 40 degrees centigrade temperature, 0 to 40 parts per thousand salinity, and 0 to 1000 atmospheres pressure. For these calculations the temperature range has been extended to -1.0°C. Table 3 contains comparative values of density calculated using LaFond's tables and Eckart's formula.

TABLE 3

DENSITY CALCULATIONS USING LAFOND'S TABLES AND ECKART'S FORMULA

Depth m	Temp °C	Salinity Ppt	Tables gm/cm^3	Difference gm/cm^3	Eckart gm/cm^3
200	6.35	35.19	1.02838	.00014	1.02852
300	4.50	35.06	1.02918	.00004	1.02914
400	2.30	34.97	1.02983	.00002	1.02981
500	0.95	34.94	1.03039	.00002	1.03041
600	0.20	34.93	1.03092	.00004	1.03096
800	-0.40	34.92	1.03189	.00009	1.03198
1000	-0.62	34.92	1.03284	.00012	1.03296
1200	-0.75	34.92	1.03378	.00015	1.03393
1500	-0.87	34.92	1.03519	.00018	1.03537
2000	-0.95	34.92	1.03751	.00023	1.03774

The data for depth, temperature, and salinity are those described by Mosby (1959) as "normal" for Station Mike.[1] The differences in density at each level are apparent. According to Eckart, there is no reason to

[1]Mosby determined his "normal" distribution of temperature and salinity with depth by finding the depth, on each of more than 200 hydrographic soundings which reached 2000 meters, of the integer isotherms and isohalines. These depths were then averaged and the resulting values plotted against the standard depths. In this way, according to Mosby, truer representations of the average gradients of temperature and salinity were retained.

believe that the LaFond results are more accurate than those obtained using the Eckart formula. The significance of the differences and their relation to the calculation of stability will be discussed below.

The stability of a column of water is calculated layer by layer. When evaluating equation(7)for E_r, the average speed of sound and the average density in each layer are used. There are two ways in which these average values can be determined. First, the values of temperature, salinity, and depth for each level in a hydrographic sounding can be substituted into equations (11) and (12). The resulting values of the speed of sound and density can then be averaged and considered to represent the "average" sound speed and density within the layer. Second, the values of temperature, salinity, and depth can be averaged within each layer and the averages substituted into equations (11) and (12). These results also represent the "average" sound speed and density within the layer. Table 4 shows densities and speeds of sound calculated using the two approaches. The data are typical values of depth, temperature, and salinity for Station Mike. The columns entitled ρ_{av} and C_{av} represent the arithmetic averages of the values of ρ and C at the top and bottom of the layer. The columns headed $\rho_{tsh,av}$ and $C_{tsh,av}$ represent the calculations made with equations (11) and (12), using the average values of T, S, and h within each layer.

The question raised is one of linearity. Which are more linear with depth, the speed of sound and density or the temperature and salinity? An adequate answer to the question requires accurate continuous determinations of these parameters simultaneously.[2] For these

[2]The small discrepancies between the two methods are well within the maximum probable error due to other causes, as will be shown below.

calculations the values of temperature, salinity and depth were averaged, since computer calculations of simple arithmetic means are faster than calculations of quadratic and higher order terms.

TABLE 4

VALUES OF DENSITY IN SITU AND SPEED OF SOUND IN SEA WATER CALCULATED FROM EQUATIONS (12) and (13) USING DIFFERENT METHODS OF AVERAGING

Depth (m)	Layer (m)	T	S_{ppt}	ρ gm/cm³	ρ_{av} gm/cm³	$\rho_{tsh,av}$ gm/cm³	C m/sec	C_{av} m/sec	$C_{tsh,av}$ m/sec
0		6.30	35.240	1.02889			1476.188		
	0-10				1.02892	1.02892		1476.174	1476.174
10		6.25	35.240	1.02895			1476.160		
	10-25				1.02898	1.02898		1476.268	1476.268
25		6.24	35.240	1.02902			1476.376		
	25-50				1.02908	1.02908		1476.566	1476.565
50		6.23	35.235	1.02913			1476.755		
	50-75				1.02918	1.02919		1476.773	1476.773
75		6.14	35.210	1.02924			1476.791		
	75-100				1.02931	1.02931		1476.700	1476.700
100		5.99	35.200	1.02938			1476.609		
	100-150				1.02950	1.02950		1476.349	1476.351
150		5.66	35.160	1.02963			1476.090		
	150-200				1.02979	1.02979		1474.100	1474.122
200		4.50	35.080	1.02995			1472.120		
	200-300				1.03029	1.03029		1468.134	1468.201
300		2.25	34.981	1.02062			1464.149		
	300-400				1.03090	1.03090		1462.471	1462.488
400		1.12	34.934	1.03118			1460.793		
	400-500				1.03147	1.03147		1459.695	1459.704
500		0.26	34.930	1.03175			1458.595		
	500-600				1.03200	1.03201		1458.495	1458.497
600		-0.15	34.918	1.03226			1458.395		
	600-800				1.03276	1.03276		1459.378	1459.379
800		-0.46	34.920	1.03326			1460.361		
	800-1000				1.03375	1.03375		1461.663	1461.663
1000		-0.63	34.917	1.03424			1462.964		
	1000-1200				1.03473	1.03473		1464.339	1464.339
1200		-0.77	34.920	1.03522			1465.713		
	1200-1500				1.03593	1.03594		1468.073	1468.073
1500		-0.85	34.916	1.03665			1470.432		
	1500-2000				1.03784	1.03785		1474.496	1474.496
2000		-0.93	34.918	1.03903			1478.559		

5. The Accuracy of E_r

The accuracy of E_r depends upon the accuracy of the calculation of the speed of sound, the accuracy of depth determination and the precision with which density *in situ* and its variation with depth can be obtained. The limit of accuracy of the sound speed calculation is one meter per second. Thermometric depths are, according to Von Arx (1962), accurate to within five meters at depths less than 1000 meters and to within 0.5% of the depth at greater depths. The effect of these errors on the accuracy of E_r can readily be examined.

Assuming an absolute layer thickness error of δh and an absolute sound speed error of δc, the calculated value of E_r is

$$(E_r)_c = \frac{(c \pm \delta c)^2}{\rho g} \frac{d\rho}{dh \pm \delta h} \tag{13}$$

where $dh = h_2 - h_1$. Then the maximum error in E_r is given by

$$\Delta E_r(h,c) = (E_r)_c - E_r \doteq E_r \left\{ \frac{2\delta c}{c} + \frac{\delta h}{dh} \right\}. \tag{14}$$

Setting $\delta h = 0.005\,(h_1 + h_2)$ and $\delta c = 1$ gives

$$\Delta E_r(h,c) \doteq E_r \left\{ \frac{2}{c} + \frac{0.005(h_1 + h_2)}{h_2 - h_1} \right\}. \tag{15}$$

The maximum values for h_1 and h_2 at station Mike are 1500m and 2000m. A typical value for c is 1470 m/sec. Thus,

$$\Delta E_r(h,c) \doteq 0.04\, E_r. \tag{16}$$

This result holds only for layers below 1000m, but the order of magnitude of $\Delta E_r(h,c)$ should be the same above this level.

The foregoing indicates that errors in E_r resulting from errors in the speed of sound calculation and thermometric depth determination will not exceed about 4% at Station Mike.

Errors in the determination of density are less well known. The effect of these errors will now be examined.

Disregarding the small systematic errors in the equation of state, the density errors are given by

$$(d\rho)_c = d\rho \pm 2\,\delta\rho$$

and

$$(\rho)_c = \rho \pm \delta\rho,$$

where $\delta\rho$ represents the error in density determination. Then the error in E_r due to density determination is

$$\Delta E_r(\rho) = (E_r)_c - E_r \doteq E_r \left\{ \frac{2\,\delta\rho}{(d\rho)_c} + \frac{\delta\rho}{(\rho)_c} \right\}. \tag{17}$$

The committee on Oceanography, National Academy of Sciences (Vetter 1959) stated that specific volume determinations(and hence, to a good approximation, density determinations) can be considered accurate to better than one in 10,000. Thus, taking ρ = 1 gm/cm^3, $d\rho$ = 0.001 gm/cm^3, and $\delta\rho$ = 0.0001 gm/cm^3:

$$\Delta E_r(\rho) \doteq 0.2\, E_r. \tag{18}$$

A 20% error in E_r is possible in the stability calculations.

6. The Results of the Calculations

The stability, E_r, was calculated for each layer of the 1253 hydrographic stations taken at Station Mike during the period January 1949 through December 1953. The monthly means for E_r were then calculated. The results are shown in figures 2, 3, 4, 5, and 6.

Certain features are evident after an examination of these figures. The surface layer (0 to 25 meters) exhibits stability during the summer months and instability during the winter months except for the year 1953, and the duration of the surface instability is longer than the duration of the surface stability. The deeper waters (below about 800 meters) exhibit small stability with E_r values only slightly greater than unity throughout the year. The striking feature, when considering the bottom water formation, is the presence and persistence of an intermediate layer of stable water. In the summer season this stable layer extends from 25 meters to about 600 meters. In the winter months it is somewhat more variable, but generally is located between 300 and 700 meters, and is never less than 200 meters thick. In the spring and fall the stable layer varies between the winter and summer distributions.

The stable layer discussed above has been delineated using an E_r value of 1.20 as the dividing line between stability and instability, in keeping with the probable accuracy of E_r. That this is a reasonable choice can be further demonstrated. Recalling that E_h^* is the potential energy per unit mass and displacement, it follows that

$$\int_{h_1}^{h_2} E_h^* \, dh = \frac{v^2}{2} \qquad (19)$$

where $\vee$ is the incident velocity a water element must have in order to pass downward through the layer limited by depths h_1 and h_2. Since E_h^* and E_r are related by

$$E_h^* = \frac{g^2}{2c^2}(E_r - 1), \tag{20}$$

it follows that

$$\vee^2 = \frac{g^2}{c^2}(E_r - 1)(h_2 - h_1). \tag{21}$$

For the typical winter thickness of the stable layer, 200 meters, and speed of sound at the depth of this layer, 1468 m/sec, equation (21) yields for an E_r value of 1.20 an incident vertical velocity of about 4.5 cm/sec. This represents the downward speed which a quantum of water would have to achieve to penetrate the stable layer and reach the bottom. Such large vertical velocities are unlikely to occur.

The mean stability values have another striking feature. The mean depth of the stable layer is, in all cases, between 600 and 800 meters. Its exact location is unknown, due to the large spacing between observations in the deep hydrographic casts. However, on six different occasions during the period 1951-1953 the depth of the base of this layer decreased sharply, moving upward to about 400 meters. This represents an increase in the thickness of the bottom-water layer of roughly 400 meters.

Also evident in the mean stability patterns are quanta of water having E_r values less than unity, indicating instability. These quanta appear from time to time during the winter and spring months, but in all cases they remain above the stable layer discussed above.

The question arises concerning the stability distribution in the individual days of the period of observation. Figures 7 and 8 show the distribution of E_r for a typical winter and summer month. The upper layers exhibit more variability than the monthly mean patterns, but the intermediate and deep layers closely resemble the monthly mean pattern. The individual deep observations show that the stable layer is never less than 200 meters thick. The value of the stable-layer E_r in the individual observations falls below 1.20 only on six occasions and then to a value no smaller than 1.16. On each of these six occasions the thickness of the stable layer was at least 400 meters, which more than offsets the effect of the slight decrease in E_r in the above velocity calculation.

Examination of the individual observations during the periods when the base of the stable layer rises confirms the increase in thickness of the bottom waters. In all six cases there is a regular reduction in E_r between 500 and 800 meters, from typical values of from 1.20 to 1.30 to values 1.05 to 1.09. The reduction persists during the period of the thickness increase in the mean pattern and then returns to its normal value.

Finally, the individual observations support the presence of unstable water layers. However, due to the non-constant interval between observations, no conclusions can be drawn concerning the vertical movement of this water, except that it always appears to remain above the stable layer.

The ability of the calculations to detect stability conditions associated with Nansen's theory for bottom water formation was tested in another manner, using data from an unpublished manuscript by Gladfelter

(1964). These data were taken in the Greenland Sea, an area where bottom water is known to be formed. The vertical stability distributions at two stations are shown in Table 5.

TABLE 5

STABILITY DISTRIBUTIONS IN THE GREENLAND SEA

Layer	Station 1 74N05W E_r	Station 2 73N00W E_r
0-10	1.07	1.15
10-20	0.55	8.21
20-30	8.39	9.43
30-50	3.80	3.06
50-75	1.48	1.12
75-100	1.11	1.34
100-150	1.12	0.99
150-200	1.04	1.18
200-250	1.04	1.02
250-300	1.03	1.07
300-400	1.06	1.03
400-500	1.01	1.02
500-600	1.02	1.02
600-800	1.03	1.02
800-1000	1.01	1.04
1000-1200	1.02	0.99
1200-1500	1.01	1.01
1500-2000	1.01	1.01
2000-2500	1.00	1.01
2500-3000	1.01	1.00

The results show clearly the stability characteristics for bottom water formation. With the exception of the surface layer both columns exhibit nearly neutral stability. Gladfelter's analysis of these stations (and many more in the vicinity) indicates that this region is one of massive bottom water formation. These calculations instill confidence in the method employed here to detect areas favorable to bottom water formation.

7. Conclusions

The stability formulas developed by Pollack are useful tools for the determination of vertical stability conditions. They are particularly adaptable to calculations using digital computers and offer results which are numerically consistent with results obtained with the less tractable formula of Hesselberg and Sverdrup.

The calculation of density *in situ* and speed of sound in sea water using the formulas of Eckart and Kinsler-Frey extend the usefulness of the Pollack stability parameters in computer operation. These formulas offer accuracies comparable to those achieved through the standard approach of using hydrographic tables.

Based upon the stability calculations, bottom water was not formed, at least in the manner associated with the Nansen theory, in the vicinity of Station Mike during the period 1949-1953. Quanta of more dense and unstable water are shown by the calculations to exist in that area. However, these quanta do not penetrate the stable layer and reach the bottom. It is possible that they move laterally and downward, reaching the bottom waters in some other location. This cannot be investigated with the available data.

The thickness of the bottom water layer at Station Mike undergoes aperiodic but definite variation. This is evidenced by the decreases in depth of the base of the stable layer. According to Crease (1965) the sill depth of the Greenland-Scotland Ridge is approximately 800 meters. This is also the approximate normal depth of the base of the stable layer at Station Mike. However, during the period 1951-1953 the base of the stable layer rose to a depth of about 400 meters on

six occasions. Cooper (1955) postulates that the current structure along the Greenland-Scotland ridge is such that it permits a pulsating rather than a continuous flow of water out of the Norwegian Sea, over the sill, and down into the Atlantic Basin. While this investigation does not relate to Cooper's theory it does indicate the possibility that pulses of rather large magnitude occur in the Norwegian Sea.

8. Acknowledgements

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Figure 1

The Bathymetry of the Southeastern Norwegian Sea, after Mosby

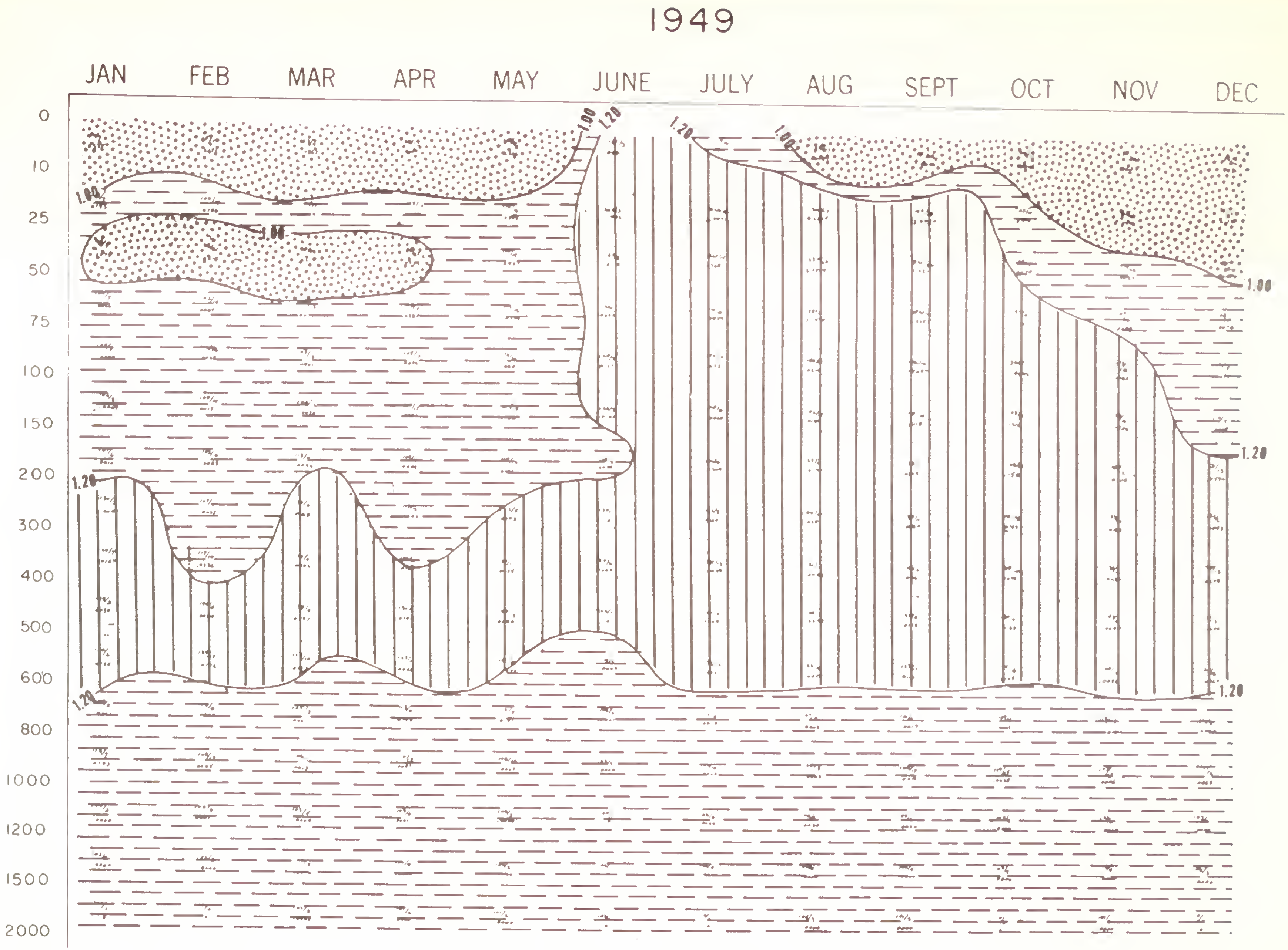

FIGURE 2

Monthly Mean Stabilities at Station Mike for 1949

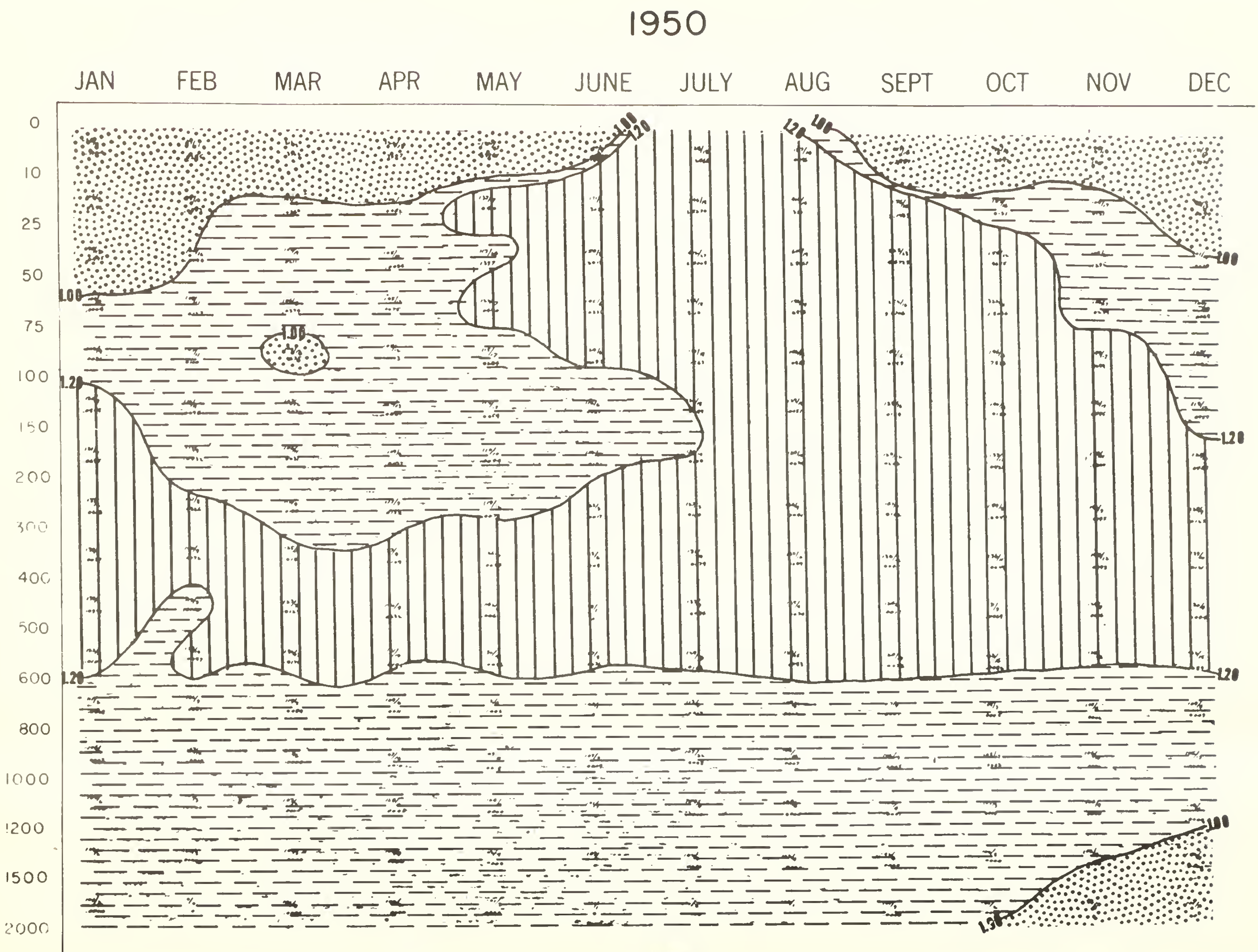

FIGURE 3

Monthly Mean Stabilities at Station Mike for 1950

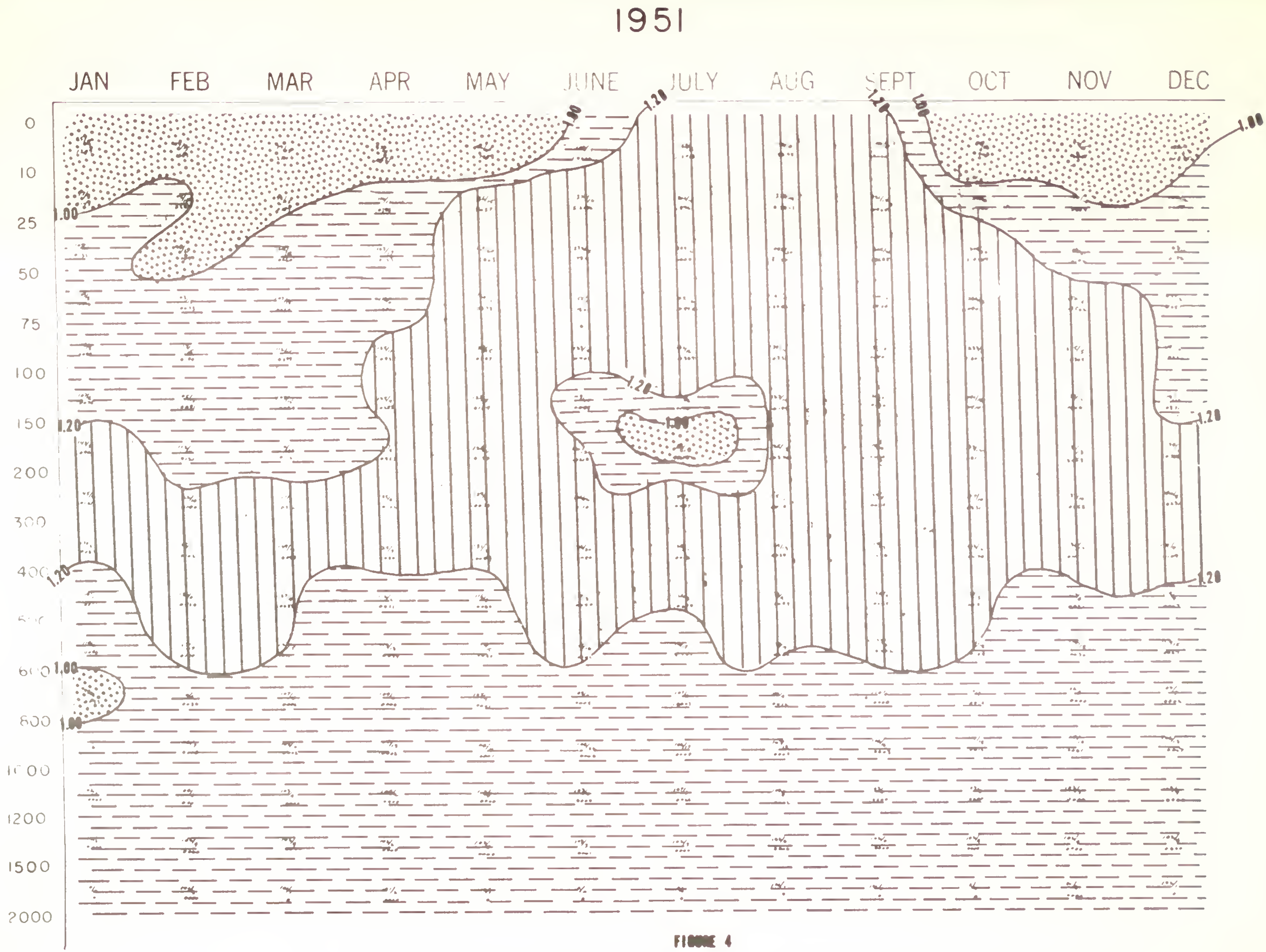

FIGURE 4

Monthly Mean Stabilities at Station Mike for 1951

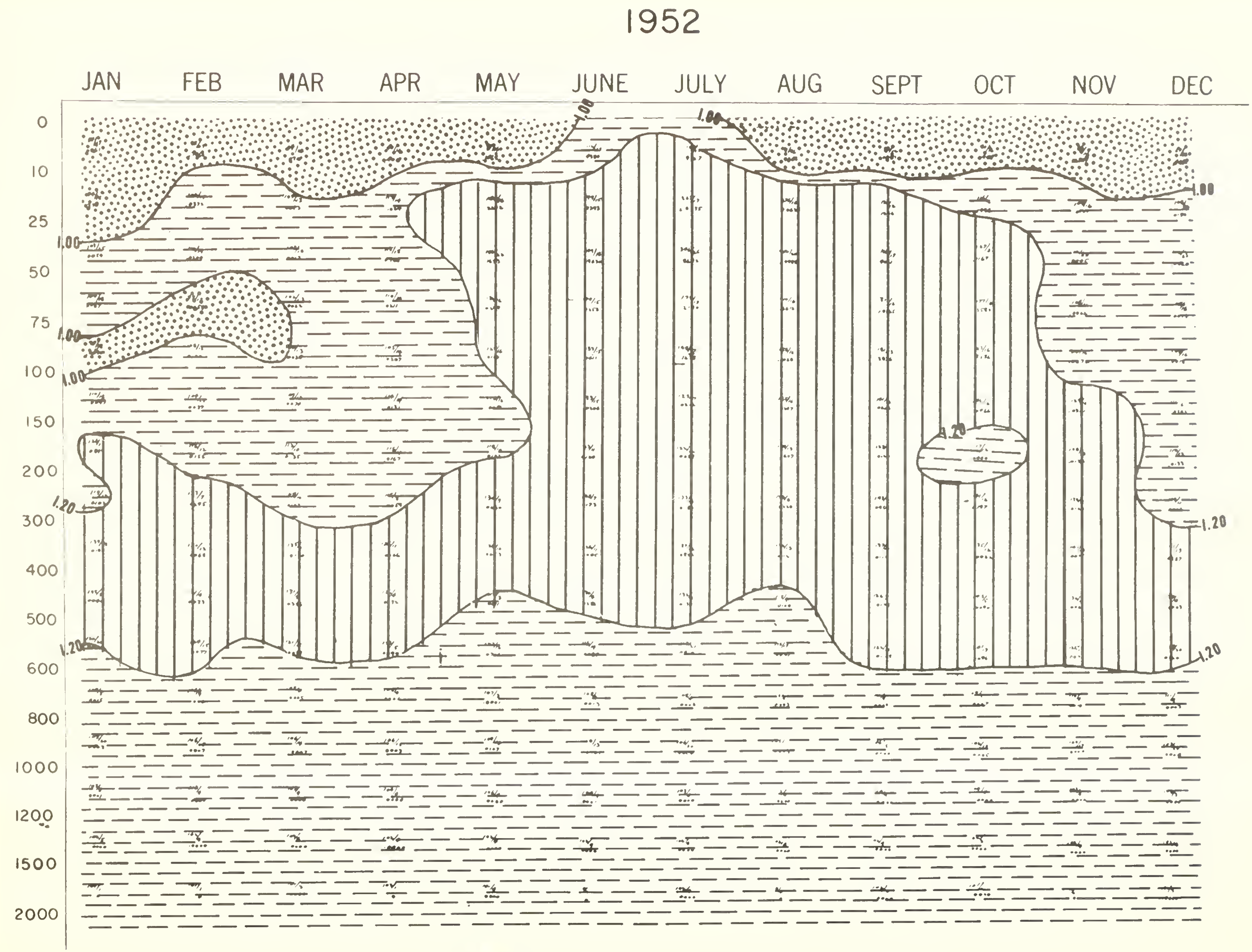

FIGURE 5

Monthly Mean Stabilities at Station Mike for 1952

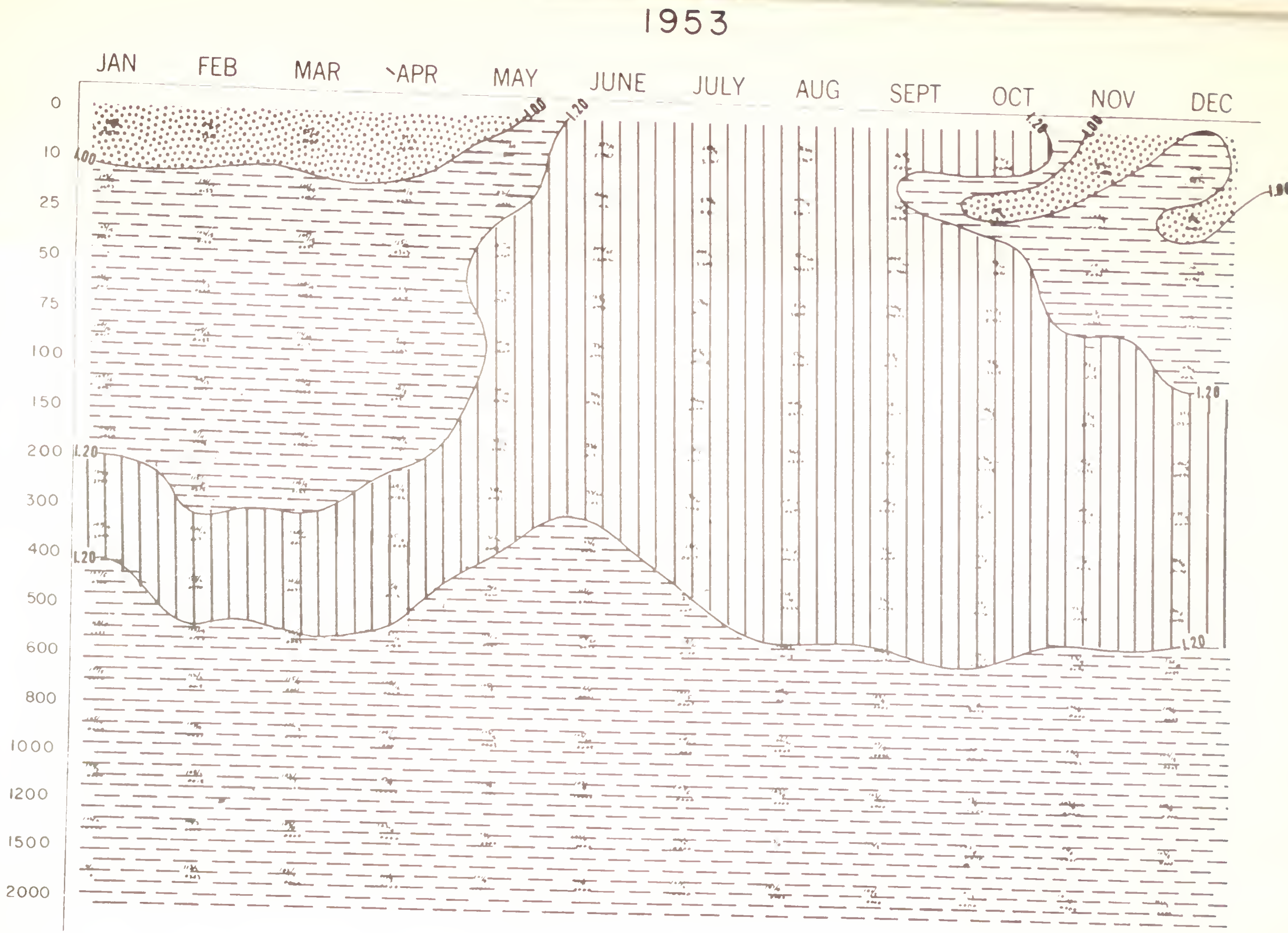

FIGURE 6

Monthly Mean Stabilities at Station Mike for 1953

FEBRUARY 1949

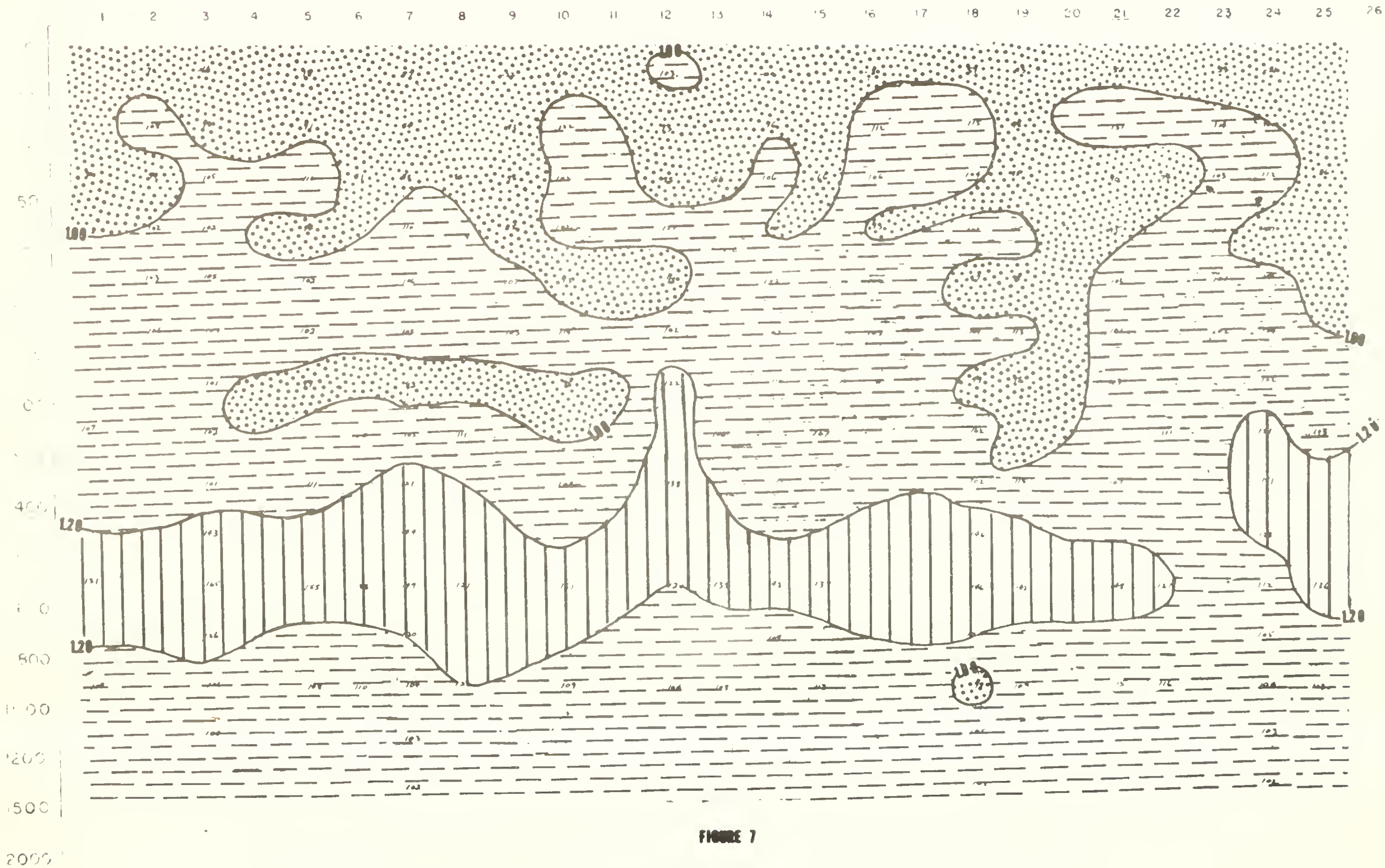

FIGURE 7

Daily Stabilities at Station Mike for February, 1949

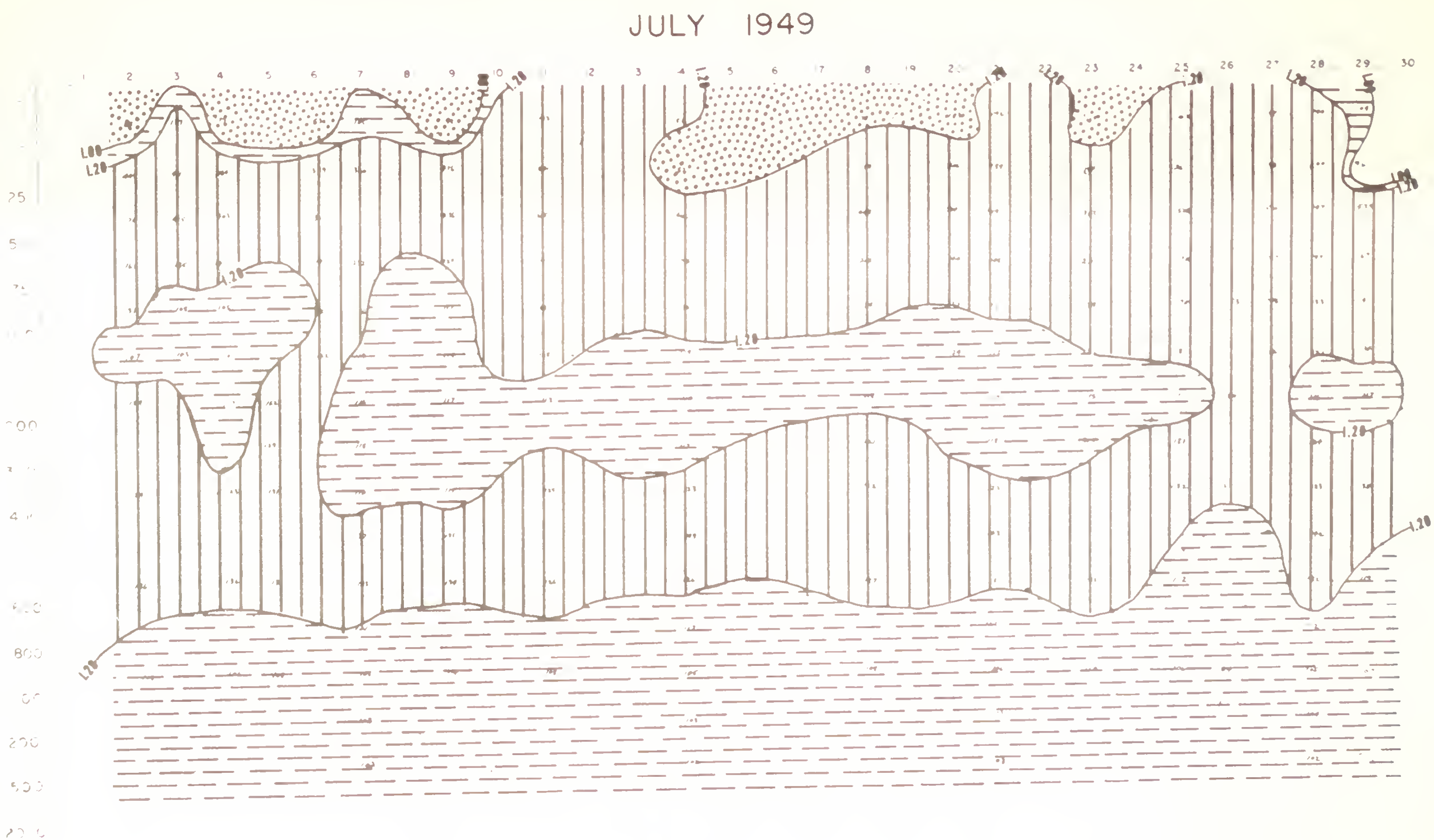

FIGURE [illegible]

Daily Stabilities at Station Mike for July, 1949

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The deep basins of the world oceans are filled with waters of very low temperature. A fundamental problem of oceanography concerns the nature of the sources of that water: their locations and temporal variations of strength. A stability analysis of temperature and salinity data recently taken in the Norwegian sea shows that bottom water probably does not form in that area.

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